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THE STRENGTH OF THE EARTH'S CRUST

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PART VIII. PHYSICAL CONDITIONS CONTROLLING THE NATURE OF LITHOSPHERE AND ASTHENOSPHERE

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INTRODUCTION AND SUMMARY

The experiments of F. D. Adams have demonstrated that under combined pressures and temperatures equal to those existing at a depth of eleven miles granite is about seven times stronger than at the surface, showing that, up to at least certain limits, the strength of the crust increases downward. The measurements of tidal deformation of the earth and of the variations of latitude concur furthermore, in proving that the rigidity of the earth as a whole is greater than that of steel. The transmission of earthquake vibrations of a transverse nature through the earth shows, not only that the earth is solid and rigid throughout, but that, on the whole, rigidity increases with depth. In none of these lines of investigation is there any clear suggestion of the existence of a thick shell of weakness—an asthenosphere.

On the other hand, the conclusion that broad areas of the crust rest in approximate isostatic equilibrium seems to imply the existence of a subcrustal zone with but little strength, readily yielding under vertical loads when these are of such breadth that the strains resulting from them are not confined and absorbed within the strong outer crust.

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[1] Section B of Part VIII, on "Relations with Other Fields of Geophysics," will be published in the succeeding number of this *Journal*.

There has thus been developed a paradox, an apparent conflict of evidence which becomes more insistent of explanation with continued accumulation of proofs of high rigidity from the domain of geophysics and of proofs of regional isostasy from the equally precise field of geodesy.

In the consideration of such broad problems, the hope of an ultimate definitive solution rests upon the use of the method of multiple working hypotheses. The surety, significance, and breadth of application of the facts must be established. By these the various hypotheses must be tested and molded. All hypotheses must be kept for further consideration provided they are not positively excluded. In the complexity of relationships there is commonly a complexity of cause, and hypotheses which seem at first to be mutually exclusive may be found to co-operate in giving a completer explanation. Those which originally appeared antagonistic may thus come to be seen as participating and dividing the field of cause between them. A paradox often points to this kind of a conclusion.

To pass to the particular problem of the relations of isostasy to the physical conditions of the earth's interior; the hypothesis developed in this study—of the existence of a zone of weakness underlying a zone of strength—must not be regarded at present as the only available hypothesis. Searching investigation must be carried forward to see if other and possibly antagonistic hypotheses cannot be developed which will equally well co-ordinate and explain the facts. Even if true in the main, it is likely, as has been the case with other hypotheses, that time will show that in certain directions it has been carried too far. Such testing, however, can best be done by others, and after the implications of this hypothesis are seen. In this part, the concluding article of this series, a discussion had best be given of the lines of adjustment by which the hypothesis here favored may be brought into harmony with other fields of geophysical evidence. With this understanding of the relation of the present investigation to the method of multiple working hypotheses, examination will be made of the paradox which has been drawn between certain conceptions from other lines of investigation and those drawn from this study of crustal strength.

Having given this introductory presentation on what is conceived to be a judicial point of view, we may turn to a review of the conclusions reached in this article. It is pointed out that rigidity is strictly a measure of stiffness; whereas a very different quality, the limit of elastic yielding, or the beginning of flow, is the measure of strength. But mass flowage may take place in a number of quite different ways, according to the nature of the solid and the environing physical and chemical conditions. The elastic limit and hence the strength will differ in the same solid according to the mode of yielding. Four modes may be here enumerated in what is thought to be their order of increasing importance, the fourth mode being that which is conceived as operative especially in the asthenosphere, and serving to maintain the condition of approximate regional isostasy.

First, flowage may take place rapidly by true plastic or molecular flow, as with lead or white-hot iron, the solid, when stressed well beyond the elastic limit, behaving like a viscous fluid. It is not thought that the terrestrial deformations are often carried on with a rapidity which requires true plastic yielding. In fact, under such rapid stresses as those produced by earthquakes and tides it is not improbable that the strength of the earth may progressively increase with depth.

As a second and quite different mode, deformation may take place by molar as distinct from molecular shear. In the zone of fracture this is manifested in jointing and faulting and is emphasized as distinct from rock flowage, but where the fracturing becomes so closely spaced as to result in slicing of individual minerals it passes under the category of granulation. Where carried on at depth there is always some degree of cementation by recrystallization. Deformation by such close-grained fracture without complete loss of cohesion is classed as rock flowage. It is thought to be developed to some degree within the lithosphere, especially by great horizontally compressive forces, but is not regarded here as the mode of deep rock flowage involved in the isostatic readjustment of unfolded tracts.

Thirdly, flowage may take place in some minerals, as calcite and ice, by gliding upon the cleavage planes. But such gliding is not regarded as the mode by which the foliated rocks are developed. It requires furthermore far greater force than that which is given by the departures from isostatic equilibrium.

A fourth mode of rock flowage is by recrystallization. It is the chief factor, as Van Hise has shown, in the deformation of the crystalline foliates. It is thought that this is also the method by which the asthenosphere yields and that a readiness of recrystallization under unbalanced stresses of a permanent nature is the cause of the weakness of the asthenosphere.

The vibratory forces transmitted as earthquake shocks and those due to tidal strain, from this standpoint, are both rapid. Under such conditions the asthenosphere could show high order of strength. It is argued that the ease of recrystallization under constant strain becomes more marked the nearer the temperature approaches to that of fusion, or to express it better from the physico-chemical standpoint, the nearer the temperature approaches to the mutual solution point of the constitutents involved. The result is that at such temperatures the rigidity may be high and not greatly different from that at low temperatures, but for permanent stresses the elastic limit becomes low. The movement of continental glaciers with a low surface gradient, accomplished by recrystallization, illustrates the condition which it would appear exists to even a higher degree within the asthenosphere.

This conclusion carries with it the idea that within the lithosphere the temperature is in general considerably below that of fusion; whereas below, in the thick zone of weakness, the temperature must lie close to, or at, that of molten rocks. For fusion there is needed, however, the energy necessary to supply the latent heat and volume expansion. Unless this energy is supplied, the asthenosphere remains solid rock, but the least accession of internal heat, or relief from external pressure, will generate a proportionate amount of magma, diffused as liquid throughout the solid. To gather into reservoirs temporarily molten, the magma must converge by rising, analogous to the draining of melted water from glaciers; uniting, as rivulets unite into rivers, and rivers discharge into lakes. No continuous lava stratum or large reservoirs of lava could, under the terms of this hypothesis, be expected to exist within the asthenosphere. Its very weakness would prevent it from acting as a containing vessel for holding large volumes of any fluid which, for any cause such as a lower specific gravity of the fluid phase, would tend to rise. The evidence of earthquake vibrations and of resistance to tidal deformation further supports the view that the asthenosphere is not a liquid or even a truly viscous zone. On the other hand, only in the lithosphere would be found the strength needed for the storage of magma in volumes until the limit of its strength as a containing vessel was reached.

Partly guided by observation upon the metamorphic rocks, partly by theories of the nature of deformation at great depths, the argument leads to conclusions on the mode of yielding within the different levels of the crust. First, the outermost zone is observed to be a zone of fracture, weak in comparison with the thick zone below. This, the second zone, is the zone of strength and yields by flowage, but flowage which is characterized by granulation as the dominant, by recrystallization as the subordinate, mode. The expenditure of energy for a given deformation is here a maximum. In the third zone, the asthenosphere, on the contrary, flowage is conceived as taking place with but little expenditure of energy, by a ready recrystallization at the temperature of primary crystallization of magmas. Those contorted granite-gneisses seen especially in the Archean rocks, which are regarded as deformed during the final stages of crystallization, exhibit locally in the outer crust the conditions which it appears may permanently prevail within the asthenosphere.

SECTION A

RELATIONS BETWEEN RIGIDITY, STRENGTH AND IGNEOUS ACTIVITY

DISTINCTIONS IN PHYSICAL PROPERTIES RELATED TO STRENGTH

Elasticity is of two natures: that of volume and that of form. The first is possessed by matter in either the gaseous, liquid, or solid state; the second is possessed by solids only and is associated with rigidity and strength. Up to a certain degree of strain known as the elastic limit, elasticity of form in the ideal solid is perfect and is expressed by the law that the change of form, or strain, is directly proportional to the load applied, or stress. This load may be maintained indefinitely and, except for a slight relaxation, the solid shows no further yielding. Upon the removal of the load there is an elastic return to the original form, but the very last stages of the recovery are slow. The elastic nature of the whole earth, in regard to both volume and form, is shown by its capacity to transmit the several kinds of earthquake vibrations. The permanence and perfection of the elasticity of form is also implied in the power of the crust to carry loads up to certain limits for times reaching into geologic periods without exhibiting progressive viscous yielding.

Beyond the elastic limit, elasticity ceases to be perfect, and a permanent change of form occurs. This relieves part of the stress and reduces the strain to within the elastic limit. The change of form may be by rupture, in which event the strength of the body is destroyed. It may be by plastic flow, in which case the strength may be increased or decreased. Wrought iron, for example, becomes somewhat stronger as a result of forging. Granite on being mashed into gneiss becomes somewhat weaker because of the development of weaker minerals, especially the micas. In the crust of the earth, except for the outer few miles, flowage takes place without probably much change in the mineral composition and consequently in the strength of the rock. Deformation will continue as long as the stress is maintained well above the elastic limit, but upon the cessation of movement there may still remain residual stresses up to the elastic limit. If the residual stresses over broad areas are small, it may be because the development of weaker structures, such as folds or zones of igneous injection, has eased the strains.

Failure by flow brings in the distinction between viscosity and plasticity. These are often used, even by physical geologists, as merely synonymous terms, but there is a real distinction which should be noted. Fluids are viscous to a small or large degree and can have no elasticity of form. Viscous flow must, however, overcome internal friction and requires time for its accomplishment.

With long time even a minute force will cause even a very viscous fluid to flow. Solids, on the other hand, possess elasticity of form, and below the elastic limit can hold shearing stresses indefinitely. Above it they may flow and in so doing exhibit plasticity. The phenomenon differs from viscosity in that the force must rise to a certain magnitude before any gliding between molecules begins. The crust, then, is plastic but not viscous.

Although the theoretical distinction between plasticity and viscosity is clear, recognition must be given to conditions where the two states merge. This is especially true for undercooled glasses. A glass in its molecular organization is a liquid and yet it possesses definite elastic moduli and elastic limit. From this standpoint of elasticity the glass, therefore, is a solid. Upon rise of temperature there is, however, no absorption of latent heat to mark a change of state, the elastic limit gradually lowers, disappears for prolonged stresses, and elastically, the substance passes by gradation from a solid to a liquid. The existence of these transition cases should not, however, be permitted to obscure the real distinctions between solid and liquid.

The crust yields as a plastic solid to forces which strain it beyond its elastic limit. But the solid flowage which this implies may be either by distortion of crystals or by recrystallization. The first is familiar for rapidly applied forces, requires comparatively great stress, and corresponds to the usual conception of plasticity. The crystalline rocks make us familiar, however, with the idea of mass plasticity by recrystallization. This is plasticity, but in a somewhat different sense from that which is usually conveyed by the term.

The degree of elasticity which a substance may exhibit is a different property from the elastic limit. A bar of wrought iron one square inch in cross-section will be elongated one part in 28,000,000 by a tensile stress of one pound. A similar bar of glass would be elongated one part in 10,500,000, more than twice as much. These ratios measure the degree of elasticity under tensile or compressive stresses and differ for each substance. The figure is known as Young's modulus of elasticity. A substance may be highly elastic, that is, have a high modulus of elasticity, as cast iron, or glass, and yet be brittle because of a low elastic limit under rapid tensile stress, combined with lack of plasticity at ordinary temperatures. At temperatures sufficiently high, the modulus is not greatly different, but the elastic limit is still lower. The substance is now, however, plastic, rather than brittle, since plasticity is greatly increased; but a rapid strain, exceeding the rapidity with which plastic deformation can take place, may still produce fracture. Another substance, such as rubber, may have a low modulus of elasticity and yet a relatively high elastic limit.

Among similar substances under similar physical conditions there is, however, a definite association of these properties which for the metals is brought out well in a tabulation by Johnston and L. H. Adams.[1] It is shown for a class of substances, such as the metals, that the modulus of elasticity, the hardness, the tensile strength, and the elastic limit all, so far as the data are given, occur in the same order; so that of two metals that which has the higher elastic limit is the higher also in the other qualities. From this association there results a ready mental confusion between rigidity and strength. The one, however, denotes the degree of resistance to distortion from a unit-shearing stress and gives the modulus of ridigity. The other is measured by the elastic limit. As an example of the confusion between these two different properties, it is known that the earth as a whole is more rigid than steel. This to many would appear to mean that it was stronger than steel. Earthquake waves show that the earth becomes progressively more incompressible and more rigid with depth. This might be held as evidence against the existence of a thick sphere of weakness, the asthenosphere. High incompressibility and high rigidity are not, however, direct testimony of strength, and it is the purpose of the next topic to show under what conditions a solid may be very rigid and yet very weak.

CONDITIONS FAVORING ASSOCIATION OF HIGH RIGIDITY WITH LOW ELASTIC LIMIT

Alpine glaciers as well as the Alpine-like margins of the Greenland ice sheet move much more rapidly in the summer than in the winter, a phenomenon to be accounted for by the rate of recrystallization. The parts of an ice crystal which are subjected to shear and compression have the melting-point lowered. They melt, discharge the strain, and refreeze. In the winter the general temperature is reduced, and a greater strain is necessary to bring the melting-point down to the lower temperature. Until local melting is produced the ice behaves like any other crystalline solid, as a substance possessing elasticity of form. Beyond that point it exhibits plasticity and behaves in some respects like a very viscous fluid. In other respects, however, it exhibits properties quite distinct from that of the usual conception of mere plastic flow, since in the testing machine, or on the walls of a crevasse, ice will resist strong shearing strains, and yet the glacier as a whole yields and flows slowly under a moderate pressure-difference as shown by the low gradient of its upper surface. Glacial motion appears to take place, therefore, by the solution and growth of crystals, not by a true viscous flow. The solid and crystalline nature throughout as opposed to viscous fluidity is furthermore shown, as Chamberlin has noted, in the power of the glacial ice to shove over and abrade its floor and to ride up slopes. Chamberlin adds that a dry glacier is a rigid glacier. A dry glacier is necessarily cold, and a cold glacier is necessarily dry.[1]

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[1] "On the Effect of High Pressures on the Physical and Chemical Behavior of Solids," *Amer. Jour. Sci.*, XXXV (1913), 220.

With ice subjected to slowly applied forces the elastic limit is consequently dependent upon the point of yielding by recrystallization. We thus see an intimate relationship between temperature and variation in the elastic limit, the elastic limit for ice being greater for low temperatures than for high temperatures. But the modulus of rigidity, on the contrary, measures the elastic change of form for unit-shearing force, change of form not accompanied by crystallization, but marked by a capacity to spring back to the original form upon the removal of the stress.

T. W. Richards, in his studies on the compressibility of solids, notes that they are almost as compressible and voluminous at absolute zero as at ordinary temperatures. Under this conception the molecules, taken as equivalent to their spheres of influence, are in actual contact and suffer mutual compression owing to the attraction of cohesion. The influence of heat is relatively unimportant in determining the density of a solid. The atoms are in most cases even more compressed and distorted by the converging force of chemical affinity than are the molecules by cohesion. This corresponds with the fact that substances of small atomic volume are on the whole more incompressible than those of greater atomic volume. For the more incompressible substances also the decrease in compressibility with added load is relatively little, suggesting that they are already greatly compressed by the forces of chemical affinity.[1] The effect of heat serves only to distend slightly the spheres of influence of the atoms, so long as the substance is in the solid state. Rise of temperature to near the melting-point, as long as there is not a softening by the development of incipient liquidity, should, according to these views, change the elastic properties but slightly. The greatly lessened strength of ice near the melting-point, as expressed in the freedom of regelation, is not, following these ideas, connected with the slightly lessened incompressibility and rigidity. For many substances the problem is complicated, however, by changes of molecular state with changes in temperature and pressure. This is especially true of ice when subjected to extreme ranges in temperature and pressure, as has been shown by Bridgeman; for ordinary glacial ice, however, we deal with but a single state.

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[1] "A Contribution to the Theory of Glacial Motion," *Decennial Publications of the University of Chicago*, IX, 203, 204 (1904); Chamberlin and Salisbury, *Geology*, I (1904), 305.

For ice at a temperature of $-7^{\circ}03$ C. the compressibility has been determined for pressures ranging approximately between 100 and 500 atmospheres. It is found to possess, according to Richards and Speyers, about one-fourth of the compressibility of water at neighboring temperatures and about five times the compressibility of glass.[2] But glass possesses a compressibility between that of acidic and basic holocrystalline igneoüs rocks. Ice may be taken then as about three or four times as compressible as granite.

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[1] "The Present Aspect of the Hypothesis of Compressible Atoms," *Am. Chem. Soc. Jour.*, XXXVI (1914), 2417–39.

[2] T. W. Richards and C. L. Speyers, "The Compressibility of Ice," *Am. Chem. Soc. Jour.*, XXXVI (1914), 491–94.

Now the modulus of rigidity is related to the modulus of compressibility by means of a formula involving Poisson's ratio.[1] This ratio varies for each substance, but for rocks, for iron and steel, and probably for ice, it lies between 0.2 and 0.3 in value, so that in general the rigidity of these substances can be judged roughly by their degree of incompressibility. Consequently it is seen that glacier ice at temperatures such as those which prevail in the body of the moving glacier possesses a degree of incompressibility and rigidity which, if these elastic constants were measures of its strength, would make it wholly incapable of motion on such gradients as are observed. This can be made more obvious by some quantitative statements. Granite and similar rocks, for example, can stand permanently in steep cliffs to heights of thousands of feet. They constitute mountain ranges whose height and steepness are limited entirely by the forces of erosion on the one hand and the strength of the asthenosphere on the other. The cliffs could be very much higher and the mountains much more lofty before glacier-like flow at the base of the mountain mass would occur. In fact, with a compressive strength of 25,000 pounds per square inch, a rectangular block of granite could stand as a vertical wall 22,000 feet high, and of indefinite breadth, without yielding of the base. With a sloping face and supported by spurs such as occur in nature, the height of the granite mass could become considerably greater. For parallel mountain ranges of harmonic form and gentle slopes resting upon a foundation whose compressive strength to indefinite depths was 25,000 pounds per square inch, the mountain crests could stand eleven miles above the valley bottoms before the maximum stress-difference would reach this limit. Even then, if the slopes were as low as those of a continental ice sheet, the failure would not take place by flowage of the mountains laterally into the valleys, but by a vertical settling of the mountains and a vertical upwarping of the valleys. The lateral, plastic flow would be at some depth in the earth. If the asthenosphere were indefinitely rigid, granite mountains of sufficiently gentle slope could rise to indefinite heights. This is because the depth of maximum stress-difference would lie at about one-sixth of the wave-length below the mean level of the surface. With increasing wave-length the height of the waves could accordingly be greater without increasing the stress-difference at the trough-line of the waves. The gradient would, however, have to become more gentle; in other words, the amplitude would have to increase at a lesser rate than the wave-length. If the strength of ice were measured by its rigidity it could stand permanently in masses one-third or one-fourth as steep and high as these theoretic limits for granite mountains, without failure by plastic flow. Yet, on the contrary, the great ice fields spread out by flowage of their bases, although their surfaces possess very gentle gradients. The distinction between strength and rigidity in the movement of glaciers is thus clear. The strength of glaciers is limited by the amount of the stress-differences needed to produce slow movement by recrystallization.

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[1] Let P = Poisson's ratio; C, the modulus of rigidity; D, the modulus of compressibility. Then $C=\frac{3}{2}\ \frac{1-2P}{1+P}D$.

Johnston and L. H. Adams have applied this theory of yielding, well known as an explanation of glacial motion, to all plastic flow, and argue that even for those substances, such as the metals and rocks, in which cubic compressibility raises the melting-point, shear greatly lowers it for the parts under stress.[1] They argue from a physico-chemical basis that the most plausible explanation for flow in metals is that the shearing strain is great enough on individual points to produce a change of phase of individual molecules from solid to liquid, even at ordinary temperatures.

Apart from theory as to its explanation, the phenomenon of welding of iron shows for high temperatures a low elastic limit and ready passage beyond into plastic flow. For iron and steel, furthermore, the influence of temperature upon the rigidity has been investigated. Pisati gives the following equations in which n is the value of the modulus of rigidity for temperature t.[2] For iron—

$$n_t = 811 \times 10^6 (1 - .000,206t - .000,000,19t^2 + .000,000,001,1t^3),$$

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[1] "On the Effect of High Pressures on the Physical and Chemical Behavior of Solids," *Am. Jour. Sci.*, XXXV (1914), 205–53.

[2] *Smithsonian Physical Tables* (1904), p. 76.

for steel—

$$n_t = 829 \times 10^6 (1 - .000{,}187t - .000{,}000{,}59t^2 + .000{,}000{,}000{,}9t^3).$$

This equation for iron gives a minimum modulus at 314° C. equal to 95 per cent of the modulus at 0° C. For steel the minimum value occurs at 342° C. and is 90 per cent of the modulus at zero. At 528 °C. iron has the same modulus as at 0° C. and at 890° C. steel has the same modulus as at zero. Doubtless 1000° C. is above the limits of the data from which these formulas were derived. For this temperature they may consequently give inaccurate results, but it is of interest to note that the curve gives a modulus of rigidity for iron at that temperature 1.7 times that at 0° C. and for steel 1.1 times that at 0° C. The extrapolation prevents attaching quantitative value to these figures, but the qualitative conclusion may be reached that iron and steel at high temperatures do not exhibit less rigidity than they possess at lower temperatures. It is obvious, however, that above a certain temperature the elastic limit becomes very low, as shown by the capacity for forging, and for strains beyond this limit deformation takes place by plastic flow. That it is not merely incipient fusion is suggested by the maintenance of a crystalline condition through the process of deformation. The subject for iron is doubtless complicated by the fact that iron passes through more than one solid molecular state in being heated up to fusion. Presumably then the equation given for the relation of rigidity to temperature can only be a first approximation to the actual changes.

Let the attention be given next to the crystalline rocks which were once deep-seated and, owing to subjacent batholithic invasion, attained their crystallization at exalted temperatures. It is observed that, although the rock masses have been extensively deformed, the individual crystals have regrown during the process so as to possess compact boundaries, and an internal constitution nearly free from strain. The interpretation is that the deformations due to geologic forces were so slow and the rocks were so saturated with crystallizing agents at high temperatures that recrystallization could nearly keep pace with the deformation, even for temperatures below the range of plasticity. As understood by the students of anamorphism, the process has depended upon a readier solution of the molecules under shearing stress than of those free from such stress—solution carried on by means of the relatively minute proportion of gaseous crystallizers which were present through the rock mass. Such crystallizers doubtless facilitate the process. They form, in fact, solutions with the rock which may be regarded as mixtures with very low fusion points. But, theoretically, as the temperatures approach those of general fusion the need of such crystallizers diminishes. Moderate shearing stresses can thus liquefy the parts upon which they act and a process analogous to glacial motion sets in for solid rock. The strength of highly heated rock appears then dependent upon the amount by which the temperature is below the melting-point. That zone of the earth which is very weak may then be regarded as approximately at the temperature of fusion. To transform the solid into liquid there is needed only the energy required for latent heat and increase of volume. The proportion of liquid which is generated will vary directly with the amount of heat supplied or the amount of hydrostatic pressure removed. Magma, consequently, can be generated in this zone more readily than above in the zone of strength; but it will not be in reservoirs; rather will it be in its place of origin disseminated through the rock mass like water standing in a porous sandstone.[1]

ANALOGIES BETWEEN ASTHENOSPHERIC ROCK AND GLACIAL ICE

The theory of the asthenosphere as here presented is seen to have important relations to other branches of geology. The zone of weakness becomes especially the generator of magmas; the pyrosphere has its roots in the asthenosphere. But in the attempt to frame a logical picture of the processes which determine the ascent of magmas, the question arises how diffused liquid matter is drained away, rising and uniting at higher levels into magma reservoirs, temporarily molten, and the direct source of the igneous activity exhibited within the outermost crust. A deductive picture is as follows—one whose truth cannot be tested directly, but only by its general agreement with our understanding.

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[1] Recently the writer has learned from Mr. Bailey Willis of an unpublished paper which he gave some years ago to the Geological Society of Washington, in which he outlined his views of the nature of crustal thrusts as illustrated by the Appalachians. In that, and more recently, as a result of studies in the Alps and Andes, he has come to entertain the view that the zone of compensation, the lithosphere, shears over the zone below through the agency of molecular or mass fusion. Deep-seated horizontal shear and igneous intrusion he thus holds have important associations with orogenic movements. We have thus arrived independently at somewhat the same view of the nature of the zone of weakness. The part which recrystallization may play in promoting such movement is suggested in his *Research in China*, Vol. II (1907), "Systematic Geology," pp. 130, 131.

At the place of origin, liquid of an andesitic or basaltic nature will come to honeycomb the rock. The content of gases is presumably sufficiently high to reduce the viscosity. The liquid will then become able to transmit hydrostatic pressures, and, although comprising only a part of the rock mass, will constitute a continuous column of considerable height. Then becomes possible the second stage, the draining upward and the convergence of the fluid rock. Gravity is the ultimate cause, as in the downward movement of waters, but here the fluid, being lighter than its surroundings, tends to move upward. An explanation of this draining process has been given by Lane.[1] In a gas-saturated rock an excess of gas, or liquid and gas, would have the power of opening fissures at any depth in the zone of flow without the necessity for the existence of any tensile stress in the walls, or competence in the walls to maintain an open cavity. All that is necessary is that the excess pressure in the rising wedge of gas should be stronger than the cohesion of the rock. The fissure becomes filled with gas and fluid of lesser density than the solid rock of the walls. Consequently, the pressure transmitted from below is greater than the resisting pressure in the walls. This insinuating power, owing to the hydrostatic head due to the lesser gravity of the wedge, becomes greater the higher the wedge rises above the source, until near the surface the action may become violent and rapid. Daly also has outlined a theory of mechanism for the injection of abyssal wedges of magma into the upper crust.[2]

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[1] "Geologic Activity of the Earth's Originally Absorbed Gases," *Geological Society of America Bull.*, V (1894), 259–80.

[2] *Am. Jour. Sci.*, XXII (1906), 195–216; *Igneous Rocks and Their Origin* (1914), chap. ix.

His theory is constructed however for action within the roof of a magmatic substratum. Daly postulates a zone of tension, but the mechanism suggested by Lane does not require this and would seem to apply better to the region of generation of magmas, for there the cubic compression is enormous, it is not a zone which has been subjected to cooling, and therefore it is difficult to conceive of the cause of a system of tensile stresses within the asthenosphere.

We are now prepared to draw a closer analogy between the physical conditions of the asthenosphere and those of a glacier, noting the likenesses and also the unlikenesses. In the summer, in the case of Alpine glaciers, heat is supplied to the surface of the glacier until it is warmed to the melting-point, and part of the ice absorbs the amount required by the change of state and passes into water. This trickles along the surface until a fissure is met and the water sinks by force of gravity toward the bottom. Near the snout of the glacier the temperature of this deeper part may thus be more dependent upon this convection than upon direct conduction through the ice. The winter freezing tends to chill the deeper ice and slow its motion, but during the summer the descending water tends to raise the temperature toward the freezing-point. In parts of the glacier where the heating is more effective than the cooling, the waters drill channels and gather at the base of the glacier into streams, reaching finally the outer world. The descent, and gathering, and englacial flow of glacial waters is analogous to the rise and convergence of streamlets of molten rock.

In order to account for lateral mass movement within the asthenosphere, an imperfection of isostasy between continental interior and ocean basin, giving an isostatic gradient or slope toward the continental interior, seems a necessary postulate. Such a gradient is so low that it has not yet been sifted from those irregularities of mass which are owing to the strength of the outer crust, or, it may be, obscured by great compressive bowings of the crust. This isostatic gradient, the slope required to generate movement within the asthenosphere, is far lower than that of the surface of a continental ice sheet. The failures of the analogy between the asthenospheric and glacial states are then as instructive as the agreements. The glacier is thin and broad. Friction on the bottom is excessive and the motion requires more internal work. Much of its mass is permanently well below the freezing temperature. These are the factors which determine the steepness of the surface gradient. The asthenosphere by contrast should be deep and the differential motions within it necessary to satisfy isostasy would be correspondingly small. The temperature through a wide zone should be that of fusion under the hydrostatic pressures prevailing. The solid rock should be sodden with occluded gases, giving mobility to the growing fluid and ready to play their part in assisting recrystallization. Such a physical condition, as long as there is a continuous solid, would exhibit perfect elasticity and high rigidity during the passage of transverse vibrations, yet would slowly yield to prolonged shearing stresses, even though these were very small in amount.

RELATIONS OF IGNEOUS ACTIVITY TO ASTHENOSPHERE AND LITHOSPHERE

The argument has led to the view that the asthenosphere is a region where the temperature curve becomes tangent to the fusion curve, but that a condition of solidity is maintained by the recurrent elimination of that material which becomes molten. The importance of such a process, maintaining the solidity of the earth, has been dwelt upon by Chamberlin, especially as accounting for the overwhelming igneous activity of Archeozoic time. In lessened measure it applies to all later times as well.

Becker has held that the bottom of the zone of isostatic compensation is the depth at which the temperature curve approached nearest to the fusion curve, and he was the first to connect in this way the geodetic evidence with a temperature relation.[1] But Becker does not conceive of actual, permanent contact of the two curves as occurring, and took the depth of nearest approach to fusion as 122 km. This follows from Hayford's hypothesis of a uniform distribution for isostatic compensation, but in the present work there has been found reason for believing that compensation fades out through a greater depth; the strength, as measured by the existence of stress-difference, through a depth greater still. The beginning of permanent contact of the two curves, if this is the cause of the disappearance of strength, should be as much as 300 km. deep and extend through some hundreds of kilometers.

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[1] "Age of a Cooling Globe in Which the Initial Temperature Increases Directly as the Distance from the Surface," *Science*, XXVII (1908), 227–33, 392; "The Age of the Earth," *Smithsonian Miscellaneous Collections*, LVI, No. 6 (1910), 1–28.

A rectilinear projection downward of the temperature gradient observed at the surface would reach the fusion temperature of rocks at a depth of about 50 km. There must be consequently a marked curvature of the temperature gradient if the temperature and fusion curves do not meet short of 300 km. This curvature implies that near the surface there is either a greater quantity of heat flowing outward by conduction or that the conductivity of rock near the surface is very greatly decreased. But such a very great decrease in conductivity making for a higher temperature gradient finds no supporting evidence. On the other hand, a greater outward flow by conduction of heat near the surface may be due to the continued generation of heat by radioactivity to a greater degree than below; or also to a rise of magmas from the asthenosphere. Magmas which never reach the surface would bring heat by a convective process directly into the outer crust. From there the heat, slowly diffused upward by conduction, would increase the temperature gradient in the outermost part of the lithosphere. It is this factor especially which the argument of the present chapter emphasizes.

It is only within the present generation that general recognition has been given to the intrusive nature of the abyssal igneous rocks. They are now generally regarded as risen from the depths. Their action has been to break through and engulf the foundations of the ancient crust.

This process of batholithic invasion seems to be recurrent and widespread, though rise into the outer crust is restricted to the crises of diastrophism and usually reaches levels exposed to erosion only along the lines of mountain systems. The stores of heat brought up from the greater depths would be held in the crust, especially in its deeper parts, for geologic ages, blurring out in the course of time by conduction and creating a false appearance of heat lingering from an initial molten state, a resemblance increased by the added veil of new heat of radio-active origin mantling the ancient stores.

The temperature gradient under this view should naturally vary widely from place to place and from time to time. Igneous activity is the effective means by which heat is brought up from depths which on account of the slowness of conduction would be otherwise thermally isolated from the outer crust. Offset against this, cooling by conduction advances downward from the surface, dissipating not only the heat of local radio-active origin but that excess rising from the depths. The heat of the crust is not then a continually ebbing residuum from a primal molten state, but represents rather an oscillating ebb and flow, one of the balances of nature maintained through geologic time.

If this view be true—that the invasive igneous rocks have been an important factor in determining the amount and distribution of heat in the crust—it is doubtful if any sound arguments can be derived from the study of the present gradients as to the initial temperature. This conclusion is similar to the change of viewpoint in other lines of geology. It was once thought that the composition of the present atmosphere and the character of present climates were steps in a simple and continuous series of changes passing from primal conditions to a future in which the water would be absorbed into the earth and its surface transformed into a frozen desert. Now, however, it is generally recognized that since the earliest known times the surface conditions have been in a state of oscillating equilibrium. The argument of this section leads toward the view that this is true for the physical conditions within lithosphere and asthenosphere also.

[*To be concluded*]